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THE CHAMPLAIN SEA IN THE LAKE ONTARIO BASIN

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INTRODUCTION

The retreating front of the Labradorian ice sheet late in the Wisconsin glacial stage rested for a time against the northwestern slopes of the Adirondack Mountains. Between the ice front in the upper St. Lawrence Valley and the southern rim of the Ontario basin in New York state, the waters of Lake Iroquois were ponded. The history of this lake, with its outlet down the Mohawk Valley past Rome, New York, has been deciphered by Fairchild, Taylor, Spencer, Coleman, and others. Further withdrawal of the ice margin permitted the escape of the Iroquois waters through "Covey Gulf," southwest of the summit of Covey Hill, the northernmost hill on the west flanks of the Adirondacks, a mile north of the international boundary. The altitude of the Covey outlet at the present time is about 1,000 feet, while that of the Rome outlet is 460 feet, but, according to Fairchild,[1] the altitudes of the two outlets during Iroquois time were very similar, "if not practically identical" (see Fig. 1).

North from Covey Hill the land drops rapidly away to the broad, low valley of the St. Lawrence. As soon, therefore, as the edge of the ice sheet had withdrawn a mile or two farther northward, Lake Iroquois was drained. The water in the Ontario basin and the St. Lawrence Valley rapidly fell to sea-level, for the land stood at a much lower altitude then than now. Differential uplift of the Great Lakes region had commenced long before the extinction of Lake Iroquois, and it is commonly held that the Champlain Sea was then at its maximum extent. It has been more or less unconsciously assumed that the history of these sea-level waters in the upper St. Lawrence Valley was simply a slow but progressive withdrawal from the maximum of submergence to the present maximum of emergence. Data now available, however, lead to the conclusions that, when the ice barrier was removed from the St. Lawrence Valley north of Covey Hill, the level of the Champlain Sea was far below its maximum height, and that the strand line moved gradually up the valley of the St. Lawrence River and its

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[1] H. L. Fairchild, "Pleistocene Uplift of New York," *Geol. Soc. Amer. Bull.*, XXVII (1916), 235–62.

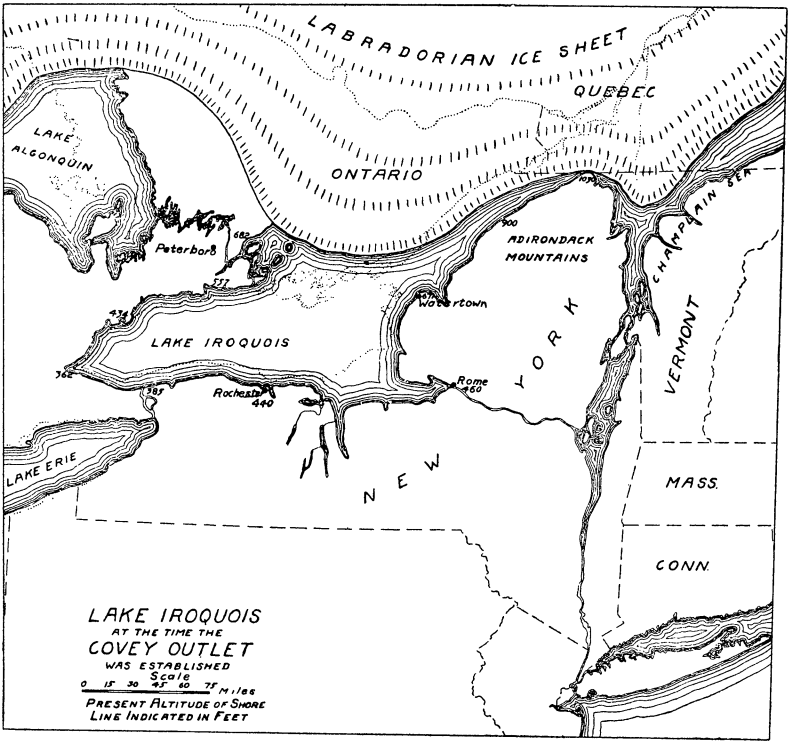

FIG. 1.—A late stage in the history of Lake Iroquois. Lake Algonquin, a portion of which is shown in the Georgian Bay region, overflows through the Fenelon Falls outlet east of Kirkfield, Ontario, and Algonquin River carries its waters to the Rice Lake embayment of Lake Iroquois. The Niagara River outflow from Lake Erie also contributes to Lake Iroquois, which is indicated in its two-outlet stage. Part of its water spills into the Mohawk Valley near Rome, New York, and part falls over the cliff at Covey Gulf, Quebec, into the marine embayment in Champlain Valley. St. Lawrence and Ottawa rivers and parts of the present Great Lakes are indicated by dotted lines. Pleistocene geography based largely upon the work of Fairchild, Coleman, Johnston, and Taylor.

tributaries to the position of greatest submergence (Fig. 2) before it began to withdraw toward its present location.

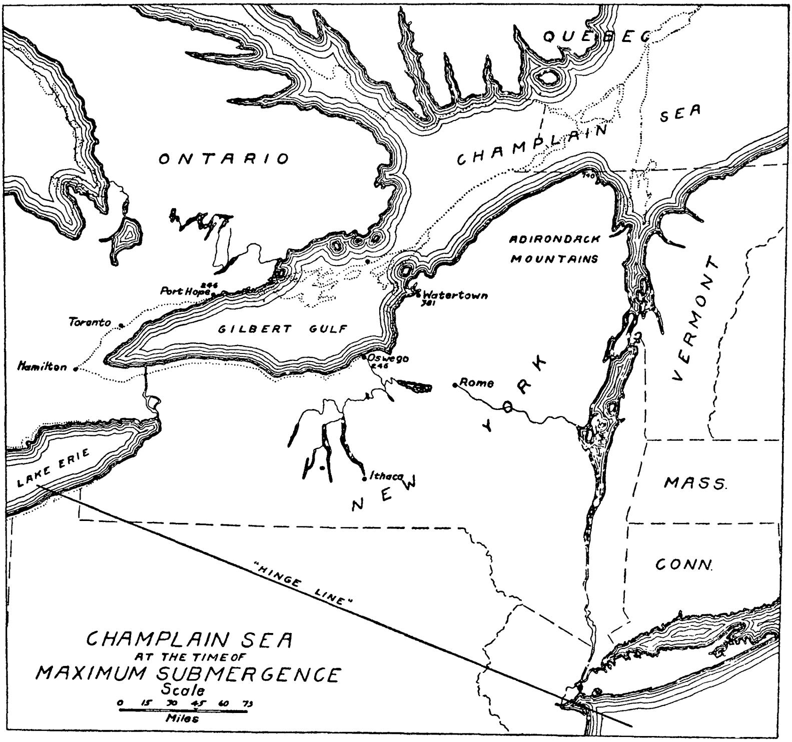

FIG. 2.—The geography of the Ontario-St. Lawrence Valley at the close of Pleistocene time. Champlain Sea is at its maximum extent. Lake Algonquin has been replaced by the Nipissing Great Lakes, which overflow northward down Mattawa and Ottawa rivers. Trent River is a comparatively small stream heading in the Trent chain of lakes and emptying into Gilbert Gulf above Trenton, Ontario. Paleogeography of New York state after Fairchild. Boundaries in Ontario and Quebec generalized and only approximately correct because of scarcity of exact data.

EVIDENCE OF A PROGRESSIVE MARINE SUBMERGENCE OF THE ONTARIO BASIN

Napanee Valley.[1]—Napanee River is one of the many southwesterly flowing streams tributary to Lake Ontario along its northern shore. The river heads in a chain of little lakes which dot the surface of the pre-Cambrian rocks in the central part of Frontenac County, 25 miles north of the east end of Lake Ontario. The lowest of the small lakes is Lake Napanee, 446 feet above sea-level. Thence the river flows across the nearly flat-lying Ordovician limestones which overlap the pre-Cambrian complex, past the towns of Yarker and Napanee to the Bay of Quinte on Lake Ontario (see Fig. 3). From Lake Napanee to Lake Ontario the length of the river is about 24 miles, and its total fall is 200 feet. Throughout this portion of its course Napanee River occupies a pre-Wisconsin youthful valley cut in the limestone cuesta and consequent upon its slope. The valley floor is 75–125 feet below the intervalley upland surface and is bounded by abrupt limestone scarps. Near Yarker the valley walls are scarcely a quarter of a mile apart, but toward Napanee they diverge gradually to a distance of a mile and a half.

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[1] I am indebted to N. B. Davis, of the Department of Mines, Ottawa, for directing my attention to the surficial deposits in Napanee Valley as well as for valuable suggestions made in the field in September, 1916.

The upland surface on either side of Napanee Valley displays abundant evidence of wave and current action. Bedded clays in many places provide a thin veneer of fertile soil over the barren glaciated limestone surface. Elsewhere, bared rock surfaces are studded with the larger bowlders from glacial drift; all the finer products of the glacial mill have been washed away by currents and waves. No distinctive shore features have as yet been observed along the Napanee Valley to mark the upper limit of wave action, but marine or lacustrine clays are present up to altitudes of at least 450 feet in the vicinity of Yarker. At Inverary, 15 miles due east from Yarker and 11 miles north from Kingston, M. B. Baker and I have definitely located the limit of wave action at elevations between 500 and 510 feet above sea-level.

The region is within the area covered by the ice barrier to which Lake Iroquois owed its existence and could not have been freed from its glacial burden until after the extinction of that lake. The wave and current action is therefore that of Gilbert Gulf, the portion of Champlain Sea which occupied the Ontario basin after removal of the ice dam from the Thousand Island region.

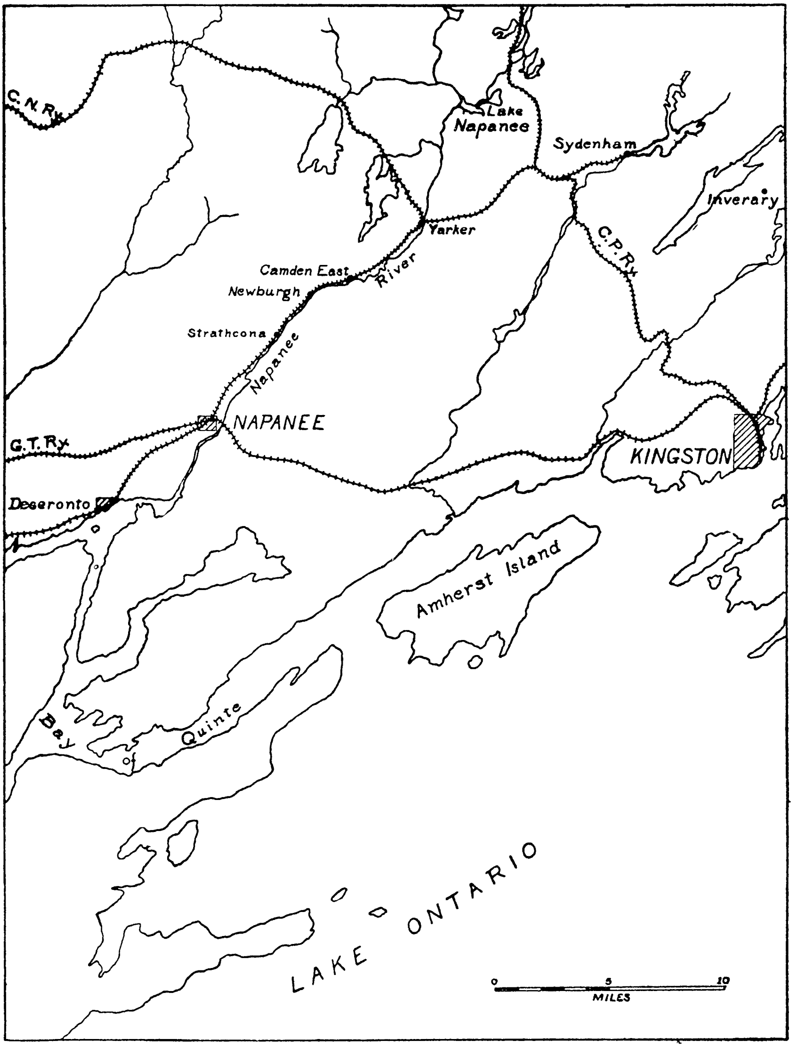

FIG. 3.—Index map of the Kingston-Napanee district, Canada, along the northern shore of the eastern end of Lake Ontario.

Near Ottawa the upper limit of marine submergence has recently been determined[1] to be 690 feet above sea-level. At Belleville the highest marine plane has an altitude of 323 feet.[2] If the gradient is regular between those two localities, the altitude at Yarker would be close to 440 feet. The nearest point on the shore line of Gilbert Gulf in New York, where its features have been mapped by Fairchild, is a small hill 4½ miles southwest from Clayton.[3] This hill is 35 miles S. 70° E. from Yarker; the Gilbert shore encircles it at an elevation slightly above 400 feet. The post-Champlain isobases in this neighborhood run approximately east and west. It is, therefore, quite clear that the Napanee Valley and adjacent uplands were submerged beneath the waters of Gilbert Gulf at least as far northward as Yarker (elevation 425 feet).

That this submergence did not take place when the ice front stood south of the divide at the head of Napanee Valley is clearly indicated by the presence of a valley train of fluvio-glacial gravels within the valley. Remnants of the gravel beds are present at many localities between Yarker and Napanee. Between Newburgh and Strathcona their fluvio-glacial, rather than glacio-lacustrine or glacio-marine, origin is readily apparent. As indicated in Fig. 4, the gravel train has a plane upper surface and its remnants are now distinctly terrace-like. The summit of the valley train is approximately 50 feet above the modern valley flat and has a similar gradient, 8 or 10 feet to the mile. The gravels and sands are irregularly bedded; cross-bedding is common, and assortment according to size of pebbles is very incomplete. Many of the stones are striated or faceted by glacial action. The valley train is in every way a typical sub-aerial fluvio-glacial deposit similar to the comparable outwash gravels of the Fox River Valley in Illinois.

The lowest and most southwesterly remnant of the Napanee valley train is within the city limits of Napanee at an elevation of less than 325 feet. In a small gravel pit beside the east abutment of the stone railway bridge the gravels and sands are overlain by bedded clays. Similar bedded clays are present on the floor of the valley eroded into the fluvio-glacial gravel deposits.

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[1] W. A. Johnston, "Late Pleistocene Oscillations of Sea-Level in the Ottawa Valley," *Canada Geol. Survey*, *Mus. Bull. 24*, 1916, p. 5.

[2] F. B. Taylor, "Gilbert Gulf," *U.S. Geol. Survey Mon. 53*, 1915, pp. 445–46; A. P. Coleman, "Marine and Freshwater Beaches of Ontario," *Geol. Soc. Amer. Bull.*, XII (1901), 129–46.

[3] H. L. Fairchild, "Pleistocene Features; Clayton-Lafargeville District," *N.Y. State Mus. Bull. 145*, 1910, pl. 46.

At the time of construction of the Napanee valley train the retreating front of the ice sheet had withdrawn less than 10 miles north of Yarker, and the marine plane could not have been over 325 feet above present sea-level. Further retreat of the ice cut off the supply of gravel, and dissection of the valley train commenced. Subsequently sea-level waters crept upward and submerged the whole region to at least the 425-foot contour line. In these waters bedded clays accumulated.

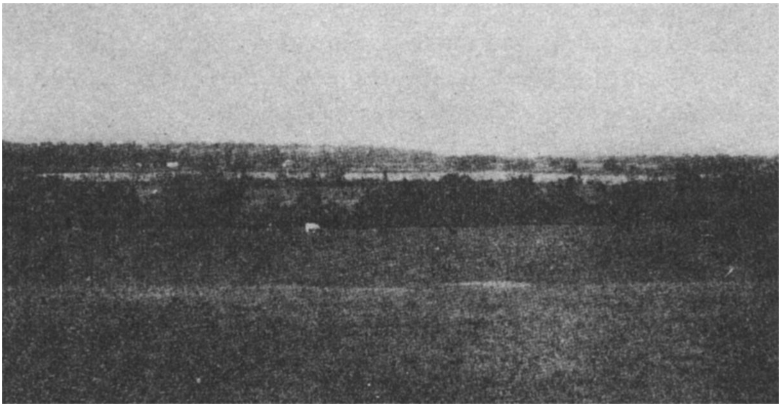

FIG. 4.—Napanee Valley between Newburgh and Strathcona, Ontario. The cattle in the foreground and the houses and barns in the distance are on the surface of the Napanee valley train. The lower flat in the middle distance is the modern floodplain of Napanee River. The exposure of fluvio-glacial gravels and sands in the escarpment overlooking the valley flat is in part due to gravel pits and in part to the railroad which parallels the river at the foot of the scarp. Beyond the buildings in the distance, the land rises abruptly to the summit of the limestone cuesta which forms the sky line at the left.

Trent Valley.—The anomalous relation between the Trent Valley spillway from Lake Algonquin and the Gilbert Gulf shore features has long been a puzzle to glacialists. The physiographic features have been described recently by Taylor[1] and Johnston[2] and need not be dwelt upon here. Briefly, Algonquin River carried the overflow from Lake Algonquin to Lake Iroquis and, after its extinction, to Gilbert Gulf. A large delta in the Rice Lake region marks the point at which the river debouched into Lake Iroquois. It is now approximately 620 feet above sea-level. The spillway channel continues down Trent Valley past the Gilbert Gulf shore line (320 feet) to the present level of Lake Ontario (246 feet) near Trenton. The channel below the summit marine plane apparently differs in no way from the portion above that level.

To explain these features by postulating a continuance of the Algonquin River flow until after the lower portion of Trent Valley had been lifted above the level of Champlain Sea is obviously difficult. Neither Taylor nor Johnston is satisfied with that explanation. In the light of the conclusions resulting from the study of Napanee Valley, the difficulties met with in Trent Valley are removed. Algonquin River carried its large volume of water to, and below, the present level of Lake Ontario *before*, rather than *after*, the Trenton neighborhood was submerged beneath sea-level waters. The delta which must have been built at the outlet of the river into Gilbert Gulf is now hidden beneath the lake. Soundings in the Bay of Quinte may some time reveal its hiding-place. By the time the Gilbert strand had crept above the present lake-level retreat of the ice in the Nipissing region had uncovered a new outlet for the upper Great Lakes, and the Lake Algonquin overflow was diverted from the Fenelon Falls outlet.

Confirmatory evidence.—Physiographic features in Napanee and Trent valleys are thus explainable by postulating a positive movement of the strand line in the Ontario basin during the final stages of the waning ice sheet. A similar advance of marine waters in Ottawa Valley has been suggested by Johnston[3] to explain the relations between the fossil-bearing clay zones in the vicinity of Ottawa.

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[1] F. B. Taylor, "The Pleistocene of Indiana and Michigan and the History of the Great Lakes" (Leverett and Taylor), *U.S. Geol. Survey Mon. 53*, 1915, pp. 445–46.

[2] W. A. Johnston, "The Trent Valley Outlet of Lake Algonquin and the Deformation of the Algonquin Water-Plane in Lake Simcoe District, Ontario," *Canada Geol. Survey Mus. Bull. 23*, 1916.

[3] W. A. Johnston, "Late Pleistocene Oscillations of Sea-Level in the Ottawa Valley," *Canada Geol. Survey Mus. Bull. 24*, 1916.

The quaternary deposits near Waterville, Maine, suggest a similar movement of the sea-level there. H. P. Little[1] states that "the main mass of the fluvio-glacial deposits is found in an esker . . . . bordered by marine clays and sands. These overlap the esker and are separated from its gravels by an unconformity considered due to sub-aerial erosion."

CHANGES IN SEA-LEVEL INCIDENT UPON THE WASTING OF THE LABRADORIAN ICE SHEET

The suggestion is an obvious one that the positive movement of the strand line in late Pleistocene times in Northwestern North America may be a result of the return of water to the ocean basins from wasting ice caps. The simple effect of ice melting is, however, complicated in the region under discussion by close proximity to the ice masses. Here, too, sea transgression has been followed by a presumably much greater negative movement of the strand, which has continued nearly or quite to the present time. Barrell's[2] discussion of the various problems involved is especially stimulating in this connection.

Three factors enter into the local problem: elevation of sea-level due to return of water which had been congealed on the land; depression of sea-level due to decreasing gravitative attraction of the ice masses; and uplift of the land due to isostatic, or other, readjustment consequent upon removal of ice burden. None of the three can be exactly evaluated from the data now available, and the effect of the third can be estimated only from field evidence. Mathematical calculations will, however, help to crystallize opinion concerning their interaction.

Woodward's classic contribution[3] to the subject does not exactly meet the situation at hand. His assumptions concerning area and thickness of the ice cap were made with the expressed purpose[1] of determining the "maximum upheaval of the water" possible as a result of glacial conditions. I have, therefore, with indispensable assistance from my wife, computed the results obtained from his formulas on the basis of quite different assumptions, which are believed to be more in accord with known data concerning the Labradorian ice sheet.

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[1] H. P. Little, "Pleistocene and Post-Pleistocene Geology of Waterville, Maine," abstract of paper presented before the Geological Society of America, December, 1916.

[2] J. Barrell, "Factors in Movements of the Strand Line and Their Results in the Pleistocene and Post-Pleistocene," *Am. Jour. Sci.* (4), XL (1915), 1–22.

[3] R. S. Woodward, "On the Form and Position of the Sea-Level," *U.S. Geol. Survey Bull. 48*, 1888.

For mathematical purposes it is fair to assume a circular ice cap with a radius of 14°, or about 966 miles, on the earth's surface and a thickness at its center of 5,000 feet. Let its surface configuration correspond to $n = 10$ in Woodward's equation 95.[2] This would give an increasingly rapid slope of the ice surface from center to border as in the following table:

	Slope in Feet per Mile of Ice Surface at Varying Distances, Expressed in Degrees, from Center of Assumed Ice Cap					
Center	4°	6°	8°	10°	12°	Border
.00	.00	.02	.34	2.53	12.95	51.43

From equation 96 it follows that such an ice cap would lower the sea-level 82.5 feet, if all its ice were formed from moisture withdrawn from the ocean. Following the method of Woodward in § 49,[3] we find that the gravitative attraction of this ice sheet would distort the level of the sea so that the disturbed surface along the border of the ice mass would be 136 feet above the undisturbed surface. The average slope of the disturbed surface within one degree of the ice border would be 0.13 feet per mile. If the ice cap were 10,000 feet in thickness at its center, the distortional effect would be twice as great as that of a 5,000-foot cap of the same diameter and surface contour.

There is every reason for believing that the development of the half-dozen ice sheets which covered parts of North America and Europe in Pleistocene times was practically synchronous. The total effect of all must, therefore, be considered in evaluating

[1] *Ibid.*, p. 18.

[2] *Ibid.*, p. 62.

[3] *Ibid.*, pp. 65–66.

resultant changes in sea-level. Daly[1] has presented the requisite data for estimating the area, ice-covered during the Pleistocene, which is now freed from its glacial burden. If the ice averaged 5,000 feet in thickness over this area, the sea-level would have been lowered about 235 feet by abstraction of water. In comparing this figure with the 136 feet of elevation at the border of the assumed Labradorian ice sheet it should be remembered that a considerable body of Pleistocene ice existed at distances of more than 45° from its border and would have exerted an attraction in the opposite direction.

It seems safe to conclude that at no time or place was the gravitative attraction of a Pleistocene ice cap powerful enough to elevate sea-water at its border sufficiently to overcome the lowering of sea-level due to withdrawal of water from the sea. This means that so far as these two factors are concerned the movement of the strand line was everywhere negative during the time of advancing ice and positive during the waning stages.[2] Near the ice borders the movement was in each case less in amount than in lower latitudes, but its direction was the same. Disregarding secular adjustment within the earth, the retreat of the Labradorian ice front from Covey Hill northward must have resulted in a progressively more extensive submergence of the St. Lawrence-Ontario Valley.

For present purposes it is unnecessary to make any inquiry into the causes of the differential uplift of northern lands which is known to have occurred since retreat of the Labradorian ice sheet began. It is, however, essential to know the time relations between the withdrawal of the ice and the readjustments which so convincingly appear to be of an isostatic nature. Did the land mass respond so quickly that secular movement entirely compensated the rising sea-level and thus maintained a stationary, or even a retreating, strand-line? Or did crustal deformation lag behind unloading of the ice burden sufficiently to permit an upward movement of the sea-level to precede the upward movement of the land?

[1] R. A. Daly, "The Glacial-Control Theory of Coral Reefs," *Proc. Amer. Acad. Arts and Sci.*, L (1915), 172.

[2] Cf. C. Schuchert, "The Problem of Continental Fracturing and Diastrophism in Oceanica," *Amer. Jour. Sci.* (4), XLIII (1916), 92, 93.

The philosophy of diastrophism is not sharply enough defined to permit an a priori answer to this question. On the one hand are many data, such as those obtained by Michelson[1] in his study of tides, which indicate quick response to external strain; on the other hand are just as many, if not more, facts, such as those concerning periodicity of orogenic episodes, which indicate the ability of the earth to delay adjustment for some time after strains begin to accumulate.

Field evidence in the Great Lakes region strongly suggests considerable lag of continental uplift behind ice removal. Post-glacial marine fossils in the Hudson's Bay drainage basin occur at elevations at least as great as 450 feet[2] and indicate uplift of that amount since the Labradorian ice sheet shrank to a diameter of less than two or three hundred miles. The highest shore line of the Champlain Sea rises toward the north at an average gradient of more than 2½ feet per mile and is essentially parallel to the Iroquois strand—sufficient proof in itself that the uplift of the St. Lawrence region, in greater part at least, lagged considerably after removal of the ice load.

The facts concerning Quaternary diastrophism in the region covered by the Labradorian ice sheet seem to be in complete agreement with the conclusion reached from field evidence in Napanee Valley, namely, that the late Wisconsin and Recent uplift did not cause the marine strand to retreat until after an appreciable interval of sea-advance had resulted from the melting of the ice.[3]

[1] A. A. Michelson, "Preliminary Results of Measurements of the Rigidity of the Earth," *Jour. Geol.*, XXII (1914), 97–130.

[2] A. P. Coleman, "Lake Ojibway; Last of the Great Glacial Lakes," *Ontario Bur. Mines, Ann. Rep.*, XVIII (1909), 284–93.

[3] In determining the amount of deformation affecting the shore lines of an inclosed body of water, such as Lake Iroquois, during its existence, the distortion of its surface by gravitative attraction should by no means be disregarded. Retreat of the ice front will, in effect, carry the inclined plane of the water surface northward and at the same time cause it to approach more closely a horizontal position. The result will be splitting of beaches north of the lake's outlet, and drowning of the older shore features south of the outlet. This apparent, though not real, warping of the basin was in many cases of sufficient amount to be of quantitative importance. One reason why the Nipissing beach, although tilted, departs but little from a true plane, while the Algonquin and earlier beaches are warped as well as tilted, is that Lake Nipissing was not a marginal glacial lake, as were its predecessors.

SUMMARY

Withdrawal from the Thousand Island region of the Labradorian ice barrier responsible for the existence of Lake Iroquois was followed by accumulation of fluvio-glacial gravels in Napanee Valley. The pro-glacial stream descended to a locality which is now less than 325 feet above sea-level before it debouched into Gilbert Gulf, an arm of the Champlain Sea. At the same time Algonquin River carried the overflow from Lake Algonquin down Trent Valley past Trenton, Ontario, to an outlet which is now beneath the waters of Lake Ontario.

The melting-back of the ice front from Covey Hill toward the Height of Land was contemporaneous with a positive movement of the strand-line which carried marine waters toward the head of the Ontario basin and drowned the Napanee and Trent valleys. The positive movement of the strand-line was followed by a negative movement, which began approximately at the time of complete disappearance of Labradorian ice and has continued nearly or quite to the present time.

Disregarding crustal movement, the waning of Pleistocene ice caps would result in a world-wide transgression of the sea as its volume was increased by the return of water temporarily abstracted to form the ice masses. In high latitudes the amount of movement of sea-level would be much less than in low, because of gravitative attraction of the ice, but everywhere the direction of movement would be the same.

Secular adjustment following removal of ice load was delayed in the Ontario-St. Lawrence region long enough to permit a stage of sea-advance before upward tilting overcame the effect of ice-melting and the stage of sea-retreat was reached.